1   **Global Ocean Surface Velocities from Drifters: Mean, Variance, ENSO Response,**

2   **and Seasonal Cycle**

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16 **ABSTRACT**

17   Global near-surface currents are calculated from satellite-tracked drogued drifter

18 velocities on a 0.5° × 0.5º latitude-longitude grid using a new methodology. Data used at

19 each grid point lie within a centered bin of set area with a shape defined by the variance

20 ellipse of current fluctuations within that bin. The time mean current, its annual

21 harmonic, semiannual harmonic, correlation with the Southern Oscillation Index (SOI),

22 spatial gradients, and residuals are estimated along with formal error bars for each

23 component. The time mean field resolves the major surface current systems of the world.

24 The magnitude of the variance reveals enhanced eddy kinetic energy in the western

25 boundary current systems, and along the Antarctic Circumpolar Current, as well as three

26 large "eddy deserts", two in the Pacific and one in the Atlantic. The SOI component is

27 largest in the western and central tropical Pacific, but can also be seen in the Indian

28 Ocean basin. Seasonal variations reveal details such as the gyre-scale shifts in the

29 convergence centers of the subtropical gyres and the seasonal evolution of tropical

30 currents and eddies in the western tropical Pacific Ocean. The results of this study are

31 available as a monthly climatology.

32

1. Introduction

Knowledge of global near-surface currents is important for a variety of uses including ship routing, search and rescue efforts, biological and chemical studies, and both hindcasts and forecasts of the transport and dispersion of floating material including plastic and oil [c.f., *McCord et al.*, 1999; *Davidson et al.*, 2009; *Yoder et al.*, 1994; *Law et al.*, 2010; *Maximenko et al.*, 2012]. Here we apply a new methodology to derive the global distribution of time-mean near-surface ocean currents, their seasonal cycle, their projection onto the Southern Oscillation Index (ENSO), and the variance of eddy fluctuations from a homogeneous data set of global near-surface current observations from the satellite-tracked drifters of the Global Drifter Program (GDP) [*Niiler*, 2001; *Lumpkin and Pazos*, 2007].

Although previous studies have derived a global mean surface velocity field from GDP drifter observations [c.f., *Maximenko et al.*, 2009], they have suffered from contamination by undiagnosed drogue loss for a significant fraction of the data [*Grodsky et al.*, 2011; *Rio et al.*, 2011]. GDP drifters have a drogue (sea anchor) centered at a depth of 15 m to reduce the downwind slip (motion with respect to the current at 15 m depth) to ~0.1% of the wind speed for winds up to 10 m s^{-1} [*Niiler et al.*, 1995] and follow the water within the mixed layer. When this drogue is lost, the downwind slip increases to ~1–1.5% of the wind speed [*Pazan and Niiler*, 2001; *Poulain et al.*, 2009; *Lumpkin et al.*, 2012]. A recent reanalysis of drogue presence has removed undrogued data from the GDP data set [*Lumpkin et al.*, 2012]. Here we set out a new method for binning and mapping the data that captures spatial and temporal variations in the data and

54 produces formal error bars for the mapped velocity components. These results are made

55 available at http://www.aoml.noaa.gov/phod/dac/dac_meanvel.php[1].

56

57 **2. Data and Methods**

58   This analysis uses quality controlled data from the GDP from 1979 through

59 December 2011, interpolated via kriging to regular 6-hour intervals [*Hansen and*

60 *Poulain*, 1996]. Velocities every 6 hours are obtained via 12-hour centered differencing

61 of the kriged positions. Recent analyses have found a significant fraction of the

62 velocities in the time period 2002—2009 previously believed to be from drogued drifters

63 were from undrogued drifters; in this study, only data from drogued drifters are used,

64 following results of a manual reevaluation of drogue presence [*Lumpkin et al*., 2012].

65 NCEP operational 6-hour winds W are interpolated to the drifter locations and a

66 downwind slip of 7×10^{-4} W [*Niiler and Paduan*, 1995] is removed from the remaining

67 drifter velocities. The resulting velocities are lowpassed with a 2-point Butterworth filter

68 with half-power cutoff at five days to remove tidal and near-inertial components of the

69 flow, then decimated to daily values for this analysis. No attempt is made to separate

70 Ekman and geostrophic velocities [e.g., *Lumpkin and Garzoli*, 2005].

71   Following *Johnson* [2001], zonal and meridional drifter speeds are mapped in

72 elliptical bins centered on a 0.5º × 0.5º longitude-latitude grid, with the ratio of semi-

73 major to semi-minor axis and axis orientation set by the variance ellipse of eddy

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[1] Note to referees: this will be done pending publication; in the interim, the climatology is available at ftp://ftp.aoml.noaa.gov/phod/pub/lumpkin/greg/meandrift_world.mat. In this file, U, V, T contain the monthly climatology of currents and drifter-measured SST (dimensions: Lat × Lon × 12, 1) with associated error bars (Lat × Lon × 12, 2). Ubar (Lat × Lon × 11, 1) contains all 11 coefficients for the model fit Az to zonal speed, with associated error bars Ubar(Lat × Lon × 11, 2), and Vbar for meridional speed. Up2bar, Vp2bar and UpVpbar are $<u'^2>$, $<v'^2>$ and $<u'v'>$, respectively.

74   fluctuations (residuals with respect to the mapped current within the bin; see below) in

75   the bin. For this study we chose bins with a set area of $\pi \times 2°$ lat. $\times 2°$ long. Hence the

76   bin area at $\pm 60°$ latitude is half that at the equator, reflecting, albeit only weakly, the

77   reduction of the Rossby radius of deformation at high latitudes [e.g. *Chelton et al.*, 1998],

78   and its influence in reducing ocean current scales at higher latitudes [e.g., *Stammer et al.*,

79   2007].

80   Within each bin, the complex vector $\boldsymbol{u}$ containing observations $u + iv$ of zonal u

81   and meridional v drifter speed can be written $\boldsymbol{u} = \boldsymbol{Az} + \boldsymbol{u}'$, where $\boldsymbol{u}'$ are residuals (eddy

82   fluctuations), $\boldsymbol{z}$ is a vector containing 11 coefficients to be determined, and matrix $\boldsymbol{A}$

83   describes the model used to map the large-scale currents:

84          $A_j = [1 \quad SOI_j \quad \sin(2\pi t_j) \quad \cos(2\pi t_j) \quad \sin(4\pi t_j) \quad \cos(4\pi t_j) \quad x_j \quad x_j^2 \quad y_j \quad y_j^2 \quad x_j y_j].$ (1)

85   Here the subscript j indicates observation j collected at time t_j (in years). The unity term

86   models the time mean. The term SOI_j is a running five-month average of the Southern

87   Oscillation Index [*Trenberth*, 1984] centered at t_j; this term models the influence of

88   ENSO, and is especially important near the equator, where drifter observations may be

89   biased towards El Niño-related periods of weak Trade Winds and associated low

90   equatorial divergence [*Johnson*, 2001]. The sinusoidal terms fit annual and semi-annual

91   harmonics [*Johnson*, 2001; *Lumpkin*, 2003]. The final five terms in A_j capture spatial

92   gradients within the bin [*Bauer et al.*, 1998; *Johnson*, 2001] and are expressed with

93   respect to the non-dimensionalized zonal distance x_j and meridional distance y_j from the

94   bin center, expressed in degrees longitude or latitude divided by $1°$.

95   The model is evaluated iteratively, starting with a circular bin, and calculating the

96   resulting variance ellipses to determine a new ellipsoidal bin and new model coefficients

97   [c.f., *Johnson*, 2001]. The calculation is iterated a total of five times for each bin to

98   ensure convergence.

99   The Gauss-Markov estimator for *z* in a given bin is

100   $$z = R_z A^{\mathrm{T}} (A R_z A^{\mathrm{T}} + R_n)^{-1} u,$$ (2)

101   [c.f., *Wunsch*, 1996; *Lumpkin*, 2003], where R_z is the *a priori* (calculated before solving

102   for *z*) covariance matrix of the unknown coefficients in *z* and R_n is the variance structure

103   of the eddy noise *u'*. For this study each of the diagonal elements of R_x are assumed to

104   be equal to the squared range of *u* and off-diagonal elements were set to zero. In

105   principle prior knowledge could guide the values of the terms in R_n; our choice to set

106   them all equal assumes that any term in (1) could potentially account for the observed

107   variance of *u* in each bin. Elements of R_n are chosen to be

108   $$R_n(\tau) = [\mathrm{std}(u)^2 + i\,\mathrm{std}(v)^2]\,\cos(\pi\tau / 2T_d)\,\exp[\,-(\,\pi\tau / 2\sqrt{2}\,T_d\,)^2]$$ (3)

109   following *Lumpkin* [2003], with $T_d = 10.33$ days, consistent with an integral eddy time

110   scale of five days. This choice assigns degrees of freedom to the observations with the

111   assumption that observations less than five days apart are not fully independent

112   [*Lumpkin*, 2003]. The *a posteriori* error covariance matrix is

113   $$P_z = R_z - R_z A^{\mathrm{T}} (A R_z A^{\mathrm{T}} + R_n)^{-1} A R_z$$ (4)

114   [c.f., *Wunsch*, 1996]. The square roots of the diagonal terms in P_z are the formal error

115   bars for the coefficients in *z*; for example, $[P_z (1,1)]^{0.5}$ is the error for time-mean *u* at the

116   bin center. The total squared error for the mapped speed *Az* is $A P_z A^{\mathrm{T}}$. This quantity can,

117   for example, be used to derive error bars on seasonal or monthly mean values. The

118   magnitudes of the coefficients *z* and the residuals *u'* are compared against the prior

119   values (square roots of the diagonal terms in R_z and R_n, respectively) in each bin. The

120 solution in a bin is rejected if $|z|$ exceeds the prior value in $\boldsymbol{R}_z$ or if more than 40% of the

121 residuals exceeds the prior value in $\boldsymbol{R}_n$.

122   The number of daily observations within each bin is highly inhomogeneous (Fig.

123 1), being a function of deployment locations and the subsequent advection pathways,

124 including regions of divergence and convergence that impact drifter residence times.

125 Regions of poor data coverage (< 50 drifter days per square degree) are located in the

126 southernmost Southern Ocean, the southeast Pacific Ocean west of Chile, the Southern

127 Ocean south of Australia, the southwest Pacific immediately north of New Zealand, the

128 western equatorial Indian Ocean, the central equatorial Atlantic Ocean, the easternmost

129 Angola Basin, and the Java Sea (much of which is shallower than 15 m). The data are

130 also not homogeneous in time: the earliest deployments (in 1979) were conducted solely

131 in the tropical Pacific Ocean as part of the Tropical Ocean Global Atmosphere study

132 [*Niiler*, 2001]. Sustained deployments began in the North Atlantic in 1989 [*Fratantoni*,

133 2001], the South Atlantic in 1993, the Indian Ocean in 1994, and the Tropical Atlantic in

134 1997 [*Lumpkin and Garzoli*, 2005]. From 1993–2002, the array averaged 500–700

135 drifters at any given time; it was increased to its current size of ~1250 drifters in late

136 2005 and has been subsequently sustained at approximately this size [*Lumpkin and*

137 *Pazos*, 2007].

138

139 **3. Results**

140   We start with a presentation of the mean speed and streamlines as well as variance

141 ellipses. We then discuss the regression of surface currents onto the SOI. Following that,

142 we present the global seasonal cycle of the surface currents. As examples of the details

143   visible in the climatology, we discuss the seasonal cycle of surface currents in the

144   western tropical Pacific Ocean and the surface divergence fields of the equatorial interior

145   ocean basins. We conclude with a comparison to an oft-used surface current product.

146

147   **3.1. Mean fields**

148   Unsmoothed mean current speeds (to indicate current magnitudes) with

149   streamlines calculated from spatially smoothed velocities (to indicate current directions

150   as well as qualitatively, the surface divergence field) clearly illustrate the prominent

151   large-scale currents (Fig. 2) such as the western boundary currents, their extensions, the

152   equatorial current systems, and parts of the Antarctic Circumpolar Current (ACC). The

153   global histogram of mean speed (not shown) peaks at 4.5–6 cm s^{-1}, with 87% of the bins

154   having a mean speed < 25 cm s^{-1} and 98% < 50 cm s^{-1}.

155

156   **3.1.1 Prominent currents**

157   The structure of the Antarctic Circumpolar Current (ACC) is resolved as strong

158   jets associated with the Southern Ocean fronts, surrounded by regions of relatively weak

159   flow [*Orsi et al.*, 1995; *Falco and Zambianci*, 2011]. The ACC bifurcates east of the

160   Drake Passage (Fig. 2), with a significant portion flowing north in the western Atlantic as

161   the Malvinas Current, perhaps the equivalent of the western boundary for this otherwise

162   zonally unbounded current [e.g. *Baker*, 1982]. After meeting the southward-flowing

163   Brazil Current at the Brazil/Malvinas Confluence, the southern Confluence jet continues

164   eastward across the South Atlantic and enters the southern Indian Ocean basin.

165   Paralleling this to the south, the southern fronts of the ACC enter the Indian Ocean sector

166   and slowly migrates southward as they flow eastward through the Indian and Pacific

167   sectors [e.g., *Orsi et al.*, 1995; *Falco and Zambianci*, 2011]. The ACC is narrower and in

168   some areas up to 10 cm s^{-1} weaker than in previous mean current fields calculated from

169   drifters prior to the recent drogue presence reassessment [*Lumpkin et al*., 2012]. The

170   ACC is not resolved in the southeast Pacific sector due to the scarcity of drogued data in

171   this region.

172   In the northern and equatorial Indian Ocean, with its strong seasonal cycle [e.g.

173   *Schott and McCreary*, 2001], discussed below, the time-mean surface currents (Fig. 2)

174   are dominated by the northward-flowing Somali western boundary current and eastward

175   equatorial Wyrtki Jet. In the Indian Ocean, the westward-flowing South Equatorial

176   Current (SEC) at 24–6.5°S is shifted further south than in the other oceans, with a

177   maximum speed of 20–24 cm s^{-1} at 12°S. The SEC is fed in part by the Indonesian

178   Throughflow [*Gordon et al.*, 1997]. It bifurcates near 16°S upon reaching Madagascar,

179   with the northern branch heading westward to Africa and then turning northward to join

180   the East Africa Current [e.g., *Swallow et al.*, 1988]. The fate of the southern branch of

181   the East Madagascar Current when it reaches the southern end of Madagascar is less clear

182   [e.g., *Lutjeharms*, 2007], but in the mean drifter climatology it appears to form a jet

183   extending west-southwest from the southern tip of Madagascar to 28°S, 40°W. This jet

184   then splits, with one branch flowing southward to join the Agulhas return current and the

185   other flowing westward to join with the southward-flowing Mozambique Current to feed

186   the Agulhas Current. The Agulhas has time-mean speeds of 60–150 cm s^{-1}, with the

187   greatest speeds found from 32.5°S, 29°E to 34.5°S, 25.5°E. The Agulhas retroflects

188   south of Africa, at 20–23°E in the mean, to feed the eastward-flowing Agulhas Return

189   Current which exhibits prominent meanders as it flows eastward to ~45°E and then east-

190   southeastward to eventually merge with the ACC. On the east side of the Indian Ocean,

191   the Leeuwin Current flows south along the west coast of Australia [*Feng et al.*, 2003],

192   with a speed of 20–35 cm s^{-1} in the latitude band 24–26.5°S, but it appears in the mean to

193   be interrupted by an anti-cyclonic feature centered at ~29°S, 111°E; northward currents

194   on the west side of this feature oppose the Leeuwin and reduce its strength. South of

195   30°S the Leeuwin exceeds 100 cm s^{-1} in several bins. The Leeuwin Current wraps

196   around the southwest corner of Australia and continues eastward along Australia's south

197   coast.

198   In the North Pacific, the Alaskan Stream [*Reed*, 1984] is apparent in the mean

199   surface velocity from the drifter data (Fig. 2) just south of Alaska and the Aleutian Island

200   chain, petering out by about 170°E after it leaves its effective western boundary, near the

201   dateline, where the island chain curves back to the northwest. The southward-flowing

202   East Kamchatka and Oyashio, the western boundary currents of the subpolar gyre, are

203   less prominent than the northward-flowing Kuroshio of the subtropical gyre, as is their

204   eastward-flowing extension [e.g. *Qu et al.*, 2001] which slows to eastward speeds <10 cm

205   s^{-1} by 170°W.

206   The equatorial Pacific currents are prominent in the drifter velocities (Fig. 2), with

207   the westward flowing North Equatorial Current (NEC) at 9–24°N with a peak speed of 20

208   cm/s at 11°N when averaged in the zonal band 150°E–140°W. The NEC extends from

209   about 120ºW to the Philippines, where it bifurcates to feed the northward-flowing

210   Kuroshio and the southward flowing Mindanao Current [e.g. *Toole et al.*, 1990]. The

211   Costa Rica Dome stands out as a small isolated anti-cyclonic circulation centered near

212 9°N, 95°W [*Fiedler*, 2002]. To the northwest, a region of anti-cyclonic circulation at 12–

213 15°N, 105–95°W may be the rectified signature of the wind-generated anti-cyclonic

214 Tehuantepec and Papagayo eddies which are generated in this region from October to

215 July [*Palacios and Bograd*, 2005]; when the mean is recalculated using only August and

216 September data (not shown), the eastward flow on the northern half of this region is not

217 present. The North Equatorial Countercurrent (NECC) extends all the way across the

218 Pacific [e.g. *Johnson et al.*, 2002]. In the Pacific the SEC is strongest at very low

219 latitudes, with a narrow, ~45 cm s^{-1} (averaged from 180–100°W) branch at 2°N and a

220 larger branch of maximum speed 30–35 cm s^{-1} centered at 4°S at 140–110°W which

221 gradually shifts northward as it flows west, reaching 2°S by the dateline. These branches

222 are separated by eastward surface flow at the equator in the central to eastern Pacific (the

223 shoaled equatorial undercurrent) [*Johnson et al.*, 2002], although this feature is not

224 readily apparent in the smoothed streamlines. The SEC is also strong along about 15°S

225 from the dateline into the Coral Sea, where it bifurcates at the east coast of Australia,

226 with part heading north towards the equator via the Solomon Sea and the New Guinea

227 Coastal Current and another portion feeding the southward flowing East Australia

228 Current [e.g., *Ganachaud et al.*, 2008] which has a peak speed of 90 cm s^{-1} at 30.5°S,

229 153.5°E.

230   In the subpolar North Atlantic Ocean, the 30–35 cm s^{-1} East Greenland Current

231 rounds Cape Farewell to become the west Greenland current with peak speeds of 45–50

232 cm s^{-1}, and then the 15–30 cm s^{-1} Labrador Current, before meeting the Gulf Stream

233 extension near Newfoundland [e.g., *Higginson et al.*, 2011]. The northward-flowing

234 Norwegian Atlantic Current is also prominent in the drifter data [e.g., *Poulin et al.*,

235   1996]. In the southwestern tropical Atlantic, a portion of the North Brazil Current

236   [*Garzoli et al.*, 2004; *Lumpkin and Garzoli*, 2005] enters the Caribbean to form the 30–50

237   cm s^{-1} Caribbean Current, which passes north of a clearly defined Columbia-Panama gyre

238   in the southern Columbia Basin [c.f. *Richardson*, 2005]. The Caribbean Current feeds

239   into the Loop Current in the Gulf of Mexico, which in turn joins the Gulf Stream flowing

240   north to Cape Hatteras, where it turns eastward to feed the North Atlantic Current [e.g.,

241   *Schmitz and McCartney*, 1993]. In the time-mean the Gulf Stream is 70–80 cm s^{-1}, with

242   speeds exceeding 90 cm s^{-1} in the Florida Current and at its separation from the coast near

243   Cape Hatteras. The tropical Atlantic circulation is dominated by a zonally elongated

244   clockwise gyre consisting of the westward northern and central branches of the SEC, the

245   northwestward North Brazil Current (NBC) and its southeastward retroflection, and the

246   eastward NECC and Guinea Current [*Lumpkin and Garzoli*, 2005]. Very large time-

247   mean speeds of 80–140 cm s^{-1} are found in the NBC, with the largest speeds located at 4–

248   7°N, 54–50°W immediately before the NBC splits into the southeastward Retroflection

249   and the westward Guyana Current. Upwelling and cross-equatorial flow of surface waters

250   evident in the mean surface drifter climatology (Fig. 2) play a role in the warm return

251   flow of the Atlantic Meridional Overturning Circulation [e.g., *Roemmich*, 1981].

252   In the South Atlantic, the southern SEC bifurcates against the South American

253   coast at 14–14.5°S to form the north-northeastward NBC and the south-southwestward

254   Brazil Current. The Brazil Current meets the north-northeastward Malvinas Current at

255   37°S, 54°W in the time-mean. The Brazil-Malvinas Confluence is a complex, energetic

256   region [e.g. *Provost et al.*, 1992, *Lumpkin and Garzoli*, 2011]; in the mean drifter

257   climatology the Malvinas Current has speeds of 30–50 cm s^{-1} against the South American

258   shelf, with comparable speeds in the southward Brazil/Malvinas Confluence jet which

259   extends south to about 45°S. The eastward flow emerging from the Confluence region is

260   strongest at 48°S, associated with the Subpolar Front, with a secondary jet at 40°S

261   associated with the Subtropical Front [*Orsi et al.*, 1995].

262

263   **3.1.2 Interior gyres, convergence, and divergence.**

264   Global streamlines converge in the centers of the subtropical gyres, associated

265   with the convergent subtropical fronts (Fig. 2). These fronts terminate in so-called

266   garbage patches in the both hemispheres of the eastern Pacific and the eastern South

267   Indian oceans [*Maximenko et al.*, 2012]. In the Atlantic the convergences are centered

268   further to the west [e.g., *Law et al.*, 2011]. In contrast, streamlines diverge from the

269   subpolar gyres of the North Pacific and North Atlantic. Subpolar divergence south of the

270   ACC, where deep water is upwelled [e.g., *Sloyan and Rintoul*, 2001], is not resolved by

271   the drifters.

272   Divergence from the equatorial Pacific and Atlantic is also extremely prominent

273   (Fig. 2), and is associated with the cold tongues of upwelled water from below [e.g.

274   *Johnson et al.*, 2001; *Grodsky and Carton*, 2002]. However, in the equatorial Indian

275   Ocean, acceleration of the flow across the equator is consistent with weaker horizontal

276   divergence (see Section 3.3) that is not as visually apparent as in the equatorial Pacific

277   and Atlantic. In fact, there is a net southward surface flow observed at the equator,

278   closing a mean cell of subducted water in the Southern Hemisphere and upwelled water

279   in the North [*Schott et al.*, 2004]. Divergence is apparent near 8°S in the western tropical

280   Indian Ocean, where the winds bring the thermocline close to the surface [e.g., *Hermes*

281   *and Reason*, 2008], as well as in the Costa Rica Dome near 9°N, 90°W [*Fiedler*, 2002].

282   Streamlines also clearly diverge from the west coasts of North and South America

283   and Africa (apart from the latitude band of the NECC, Fig. 2), where the equatorward-

284   flowing eastern boundary currents are associated with coastal upwelling [e.g. *Bograd et*

285   *al.*, 2009; *Colas et al.*, 2012; *Hutchings et al.*, 2009; *Nykjær and Van Camp*, 1994]. In

286   contrast, the streamlines do not indicate divergence along the Australian west coast,

287   where the poleward-flowing Leeuwin Current carries warm water southward, and the

288   winds and circulation inhibit upwelling [*Feng et al.*, 2003].

289

290   **3.2 Variance**

291   The mean speed of eddy fluctuations with respect to the model fit (i.e.,

292   fluctuations with respect to the time-mean, seasonal, and SOI-related variations, equal to

293   the square root of twice the eddy kinetic energy) shows large geographic variations (Fig.

294   3). These results are comparable to those from satellite altimetry [e.g., *Ducet et al.*,

295   2000], except that the surface drifter analysis can be carried to the equator as geostrophy

296   need not be assumed. Maximum eddy speeds exceeding 70 cm s^{-1} are found in the

297   Somali Current region; values > 60 cm s^{-1} are also found in the Gulf Stream and Agulhas

298   Retroflection, and > 50 cm s^{-1} also in the Loop Current, Florida Current, the Pacific

299   NECC and northern SEC at 115–130ºW (presumably associated with energetic Tropical

300   Instability Waves), Kuroshio Current, Brazil/Malvinas Confluence, Agulhas Return

301   Current, North Brazil Current Retroflection, Mozambique Channel, and the East

302   Australian Current. Enhanced eddy energy is also seen along the path of the ACC, west

303   of the Hawaiian Islands, in the Pacific Subtropical Countercurrents, and along much of

304   the NECCs of the Atlantic and Pacific basins.

305   In contrast, "eddy deserts" with time-mean eddy speeds < 10 cm s^{-1} dominate

306   three large regions: the subtropical South Atlantic, the subpolar North Pacific, and the

307   southeastern South Pacific (~110–85ºW from 40–10ºS, and extending west to 170ºW in

308   the band 45–35ºS). Much smaller eddy deserts are found in the central subtropical North

309   Atlantic, west and south of New Zealand and south of the merged Agulhas Return

310   Current and ACC immediately south of Kerguelen Island.

311   The variance ellipses for these eddy fluctuations (Fig. 3) tend to align with most

312   of the major currents such as the Gulf Stream and Kuroshio, although this tendency is

313   less apparent in the East Australia and the Agulhas Return currents. Near the coasts, the

314   ellipses tend to align with the coastlines. In the ocean interiors they are generally more

315   circular, except at low latitudes, particularly in the Pacific and Indian equatorial bands,

316   where they are zonally elongated.

317

318   **3.3 Southern Oscillation Index (SOI) and surface currents**

319   Current speeds and directions from the regression against the SOI (Fig. 4, plotted

320   for SOI = -1, corresponding to a moderate El Niño) contain the anticipated eastward (and

321   slightly equatorially convergent) surge of surface currents in the Equatorial Pacific [e.g.

322   *Johnson et al.*, 2000]. Eastward values of 10–20 cm s^{-1} are seen in the equatorial Pacific

323   (2°S–3°N) from 160°E to nearly 160°W, and along 155–110°W from 2–3°N. A weaker

324   response mirroring the 2–3°N response is seen just south of the equator in the eastern

325   tropical Pacific. East of 145°W, the response on the equator is not statistically

326 significantly different from zero due to the scarcity of observations (Fig. 1). In the

327 western tropical South Pacific there is another patch of anomalous westward flow

328 associated with an El Niño-related shift in the South Pacific Convergence Zone [e.g., *Cai*

329 *et al.*, 2012]. Curiously, the response in the Tehuantepec and Papagayo eddy region 12–

330 15°N, 105–95°W suggests a weakening of the anti-cyclonic recirculation even though

331 more frequent and stronger anti-cyclonic eddies are thought to be generated there during

332 El Niños [*Palacios and Bograd*, 2005].

333 In the Indian Ocean, El Niño-related westward speeds of ~4–8 cm s^{-1} can be

334 found in a band stretching west-northwestward across the basin from ~12°S, 100°E to

335 ~1°N, 50°E, generally increasing toward the west. There is another patch of westward

336 flow just north of the equator south of India and Sri Lanka. The large spatial extent and

337 coherent directions in these features suggest they may be robust. Other less spatially

338 coherent patterns include a slowing of the surface currents in the Gulf Stream during El

339 Niños, eastward flow in the deep eastern Bering Sea, and anomalous westward flow in

340 the Mediterranean Sea. There are isolated patches of response in other areas, but nothing

341 aside from the patterns already discussed stands out as spatially coherent.

342

343 **3.4. Seasonal variations**

344 Monthly mean ocean currents for climatological February, August, May, and

345 November (Figs. 5–6) illustrate the seasonal cycle of global surface currents. For

346 example, the eastward Wyrtki jet in the equatorial Indian Ocean is prominent in May

347 (~50 cm s^{-1}) and November (~70 cm s^{-1}) while being absent in August and reversed in

348 February [e.g. *Nagura and McPhaden*, 2010]. Likewise the Somali Current is completely

349   spun-up in August during the Southwest Monsoon, but reversed in February near the end

350   of the Northeast Monsoon [e.g. *Schott et al.*, 1990]. The Bay of Bengal shows a strong

351   seasonal cycle as well, with current directions reversing completely [e.g., *Shenoi et al.*,

352   1999]. Interior flow across the equator is largely southward in August, and northward in

353   February, with seasonally reversing surface winds driving a shallow "cross-equatorial

354   roll" with subsurface flow opposite the surface flow [e.g. *Schott et al.*, 2002]. Further to

355   the south, the surface expression of the South Equatorial Current is substantially weaker

356   in August than in other seasons.

357   The seasonality of the NECC is clearly seen in the Pacific and Atlantic basins,

358   revealing for example an intense Atlantic NECC emerging from the North Brazil Current

359   Retroflection in August while neither exists in May [e.g., *Lumpkin and Garzoli*, 2005].

360   The Pacific NECC is quite strong in August and November, but much weaker in

361   February and May [e.g. *Johnson et al.*, 2002]. The South Equatorial Countercurrent in

362   the western Pacific Ocean at 11–8°S is clearly seen in February (and also in December,

363   January and March, not shown) while being absent in other months [*Chen and Qiu*,

364   2004].

365   The gyre-scale circulations (Figs. 5–6, streamlines) also exhibit seasonal

366   migrations, which are no doubt associated with seasonal wind shifts [vis., *Risien and

367   Chelton*, 2008]. For example, the centers of the northern hemisphere subtropical gyres

368   shift to the south in February and to the north in August. In the North Pacific Ocean

369   there is a dramatic shift in the latitude of the interior recirculation gyre, from ~20°N in

370   February to ~40°N in August. This seasonal migration may have a significant

371   modulating impact on the convergence of floating material in the "garbage patch" of the

372   North Pacific, compared to calculations based on time-mean Lagrangian displacement

373   statistics [e.g., *Maximeko et al*., 2012] in which these shifts would spuriously increase the

374   effective time-mean lateral diffusivity.

375

376   **3.4.1 Tropical west Pacific surface currents**

377   Many fine-scale details of the surface circulation are difficult to see when plotted

378   at the global scale, but are resolved by the drifter observations in the climatology

379   presented here. For example, in the western Pacific (Fig. 7) the southward-flowing

380   Mindanao Current can be seen to split immediately north of the northeastern tip of the

381   Celebes year-round, with the westward branch entering the Celebes Sea, and the eastward

382   branch feeding the NECC; the cyclonic Mindanao Eddy is located between the Mindanao

383   Current and NECC throughout the year [c.f., *Arruda and Nof,* 2003]. In Boreal winter,

384   the strongly zonal NECC separates into its main branch running along ~5°N and a

385   secondary branch flowing ESE along the northwestern New Guinea coast (Fig. 7). This

386   branch reverses through Boreal spring, and by August the WNW New Guinea Coastal

387   Current (NGCC) has surfaced. At the confluence of the Mindanao Current's eastward

388   branch and the NGCC, a narrow northeastward-flowing jet develops, the eastern side of

389   which recirculates to form the Halmahera Eddy [c.f., *Arruda and Nof,* 2003] which is not

390   present in Boreal winter but is clearly visible in Boreal summer through fall. This eddy is

391   thought to play a significant role in modulating exchange between the Pacific and Indian

392   basins through the Indonesian Throughflow [*Qu et al.,* 1999].

393

394   **3.4.2. Interior equatorial surface divergence**

395   The seasonal divergence in the equatorial interior of the three ocean basins (Fig.

396   8) contrast strongly. Equatorial divergence in the Indian Ocean peaks in August with the

397   relaxation of the Wyrkti Jet; the flow is strongly convergent in November through

398   January. An off-equatorial divergence is evident in the South Indian Ocean, centered

399   near 8°S in December–April, and closer to 5°S in May–August. This feature may be

400   associated with the thermocline ridge near that latitude in the South Indian Ocean [e.g.,

401   *Hermes and Reason*, 2008]. In contrast, equatorial divergence is present year-round in

402   the central Pacific and Atlantic basins [e.g., *Johnson et al.*, 2001; *Grodsky and Carton*,

403   2002], although the magnitude displays a seasonal modulation in both basins with

404   stronger divergence in Boreal spring vs. Boreal fall (mirroring the development of the

405   cold tongue in the eastern regions of these basins). Off-equatorial convergence is seen

406   throughout the year in the central Pacific south of the equator, but is more seasonally

407   modulated to the north with a convergence maximum at 4–5°N in May–June. In the

408   central Atlantic, convergent maxima are seen at 4–8° N during May–August, and at 6–

409   4°S during June–September. These patterns of equatorial divergence and off-equatorial

410   convergence are associated with the shallow tropical overturning cells in these basins

411   [e.g. *Perez et al.*, 2010].

412

413   **4. Summary and Discussion**

414   We use a global set of drogued drifter velocity measurements to map the time-

415   mean, seasonal and Southern Oscillation Index-related components of near-surface

416   currents at high resolution. Our approach combines aspects of three previous mapping

417   efforts [*Bauer et al.*, 1998; *Johnson*, 2001; *Lumpkin*, 2003] to address spatial and

418   temporal variations in the data and to produce formal error bars for the various

419   components (mean, seasonal cycle, SOI regression) of the currents. Our analysis also

420   takes advantage of data set corrections from a recent reevaluation of drogue presence

421   [*Lumpkin et al*., 2012], one result of which is significantly reduced speeds for currents

422   such as the ACC. However, it is possible that downwind (and down-wave) slip, which

423   has not been directly measured for drogued drifters at wind speeds >10 m s^{-1} [*Niiler et*

424   *al.,* 1995], may exceed the linear slip removed in the recent data set update by Lumpkin

425   et al. [2012], and as such there may still be downwind bias in regions of strong winds.

426   Assuming a mean density of 1025 kg m^{-3}, the global distribution of total kinetic

427   energy (Fig. 9d) is equivalent to 8.8×10^{17} J of energy in the upper 30 m, in all bins with

428   more than 12 drifter days per square degree (for reference, the global total variable

429   kinetic energy of the full-depth ocean is estimated at 3.8×10^{18} J [*Wunsch*, 1998]). Mean

430   currents contribute a substantial fraction to the total kinetic energy (Fig. 9a) where one

431   would expect: in the western boundary currents and their extensions, in the North

432   Equatorial currents and the North Equatorial countercurrents of the Atlantic and Pacific

433   oceans, and in segments of the Antarctic Circumpolar Current. These are all locations

434   where the total kinetic energy is high (Fig. 9d), and – except for the equatorial currents

435   and regions of the ACC– where there is topographic confinement. However, the mean

436   currents also constitute a substantial fraction of the total kinetic energy in the South

437   Equatorial currents of all three oceans as well as in the interior of the subpolar North

438   Pacific gyre, all eddy desert regions (with the exception of the Indian Ocean SEC) where

439   the total kinetic energy is relatively low. Globally, the mean currents contribute 2.9 ×

440   10^{17} J of kinetic energy in the upper 30 m, 33% of the total estimated KE.

441   The seasonal contribution (Fig. 9c) to the total kinetic energy (Fig. 9d) is 20–40%

442   in the Pacific and Atlantic North Equatorial countercurrents. It is substantial throughout

443   much of the Indian Ocean north of about 10°S, with maxima of 50–70% at 0–5°N, 55–

444   70ºE. Another location of strong seasonal contribution is the eastern tropical South

445   Atlantic Ocean, just west of Africa: it contributes 20–30% of the total KE throughout the

446   Angola Basin, and 40–70% against the African coast. Seasonal variations represent a

447   global total of 1.1×10^{17} J of kinetic energy in the upper 30 m, 13% of the total KE.

448   The fractional contribution of SOI-related variations (not shown) is generally

449   small, exceeding 10% only in the vicinity of the equator from 165°E to 165°W. It

450   represents a global total of 7.8×10^{15} J of kinetic energy in the upper 30 m, 1% of the

451   total.

452   The eddy contribution (Fig. 9b) to the total kinetic energy (Fig. 10d) dominates

453   much of the ocean, especially the interior of the subtropical gyres and the subpolar North

454   Atlantic Gyre. The fraction of the total kinetic energy accounted for by eddies exceeds

455   50% in 71% of the bins. However, in much of this area the total kinetic energy is small.

456   EKE dominates in the well-known eddy regions such as the Gulf Stream and Kuroshio

457   recirculation gyres, Agulhas and North Brazil Current ring pathways, and the

458   Brazil/Malvinas Confluence. Eddies also make a large contribution between the North

459   Equatorial currents and the various branches of the South Equatorial currents, where

460   tropical instability waves [e.g. *Chelton et al.*, 1999] are energetic. The eddy deserts of

461   the eastern North Pacific, eastern South Atlantic, and eastern South Pacific each have

462   portions where mean currents are dominant, but also portions where the eddies are

463 dominant, with low total energy levels throughout. The EKE represents a total of 4.6 ×

464 10^{17} J of kinetic energy in the upper 30 m, 53% of the total KE.

465 The climatology of time-mean and seasonal currents described here is available at

466 http://www.aoml.noaa.gov/phod/dac/dac_meanvel.php. This climatology can be used for

467 purposes such as model validation and product evaluation. For example, there are large

468 and spatially coherent differences (Fig. 10a) between the mean velocity in the drifter

469 climatology and the January 1993–December 2003 mean velocity from the 1/3°

470 unfiltered satellite-derived Ocean Surface Current Analysis Real-time (OSCAR) product

471 [*Bonjean and Lagerloef*, 2002] including geostrophic, Ekman, and buoyancy

472 components. While inter-annual variations may play a role in some of the discrepancies,

473 the drifter climatology is faster by up to 20–25 cm s^{-1} in the Gulf Stream and Kuroshio

474 currents, and up to 35 cm s^{-1} faster in segments of the ACC and Agulhas Return Current

475 Along the northern branch of the South Equatorial Current at 1–3°N in the Pacific, the

476 drifter climatology is 20–50 cm s^{-1} faster to the west than in the OSCAR product; on the

477 equator in the Atlantic it is 20–35 cm s^{-1} faster to the west (Fig. 10). These dramatic

478 differences may be due both to smoothing and difficulties in calculating the near-

479 equatorial quasi-geostrophic flow for OSCAR.

480 Differences between this new drifter-derived climatology and products giving

481 time-mean geostrophic speed (such as the CNES-CLS09 MDT product [*Rio et al.* 2011],

482 Fig. 10b) can be used to infer the ageostrophic component of drifter motion. Although it

483 extends beyond the scope of this study, it would be interesting to remove the

484 ageostrophic motion, recalculate the distribution of variance about the time-mean speeds,

485 and compare the results to eddy kinetic energy from altimetry [c.f., *Fratantoni*, 2001].

486   Because the drifter climatology gives total near-surface currents, it can also be used to

487   improve our estimates of wind stress and turbulent heat fluxes, which depend on the

488   relative speed of the wind over the moving ocean surface [*Fairall et al*., 2003]. Ocean

489   currents have been demonstrated to affect the wind stress field and its curl and divergence

490   [*Kelly et al*., 2001; *Risien and Chelton*, 2008].

491

492

499   **References**

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698

698   **Figure Captions**

699

700   Figure 1. Number of drogued drifter observations used to map fields at each grid point

701   bin (in drifter days per square degree).

702

703   Figure 2. Mean current speeds (colors, in cm s^{-1}) from near-surface surface drifter data

704   with streamlines (black lines). Streamlines are calculated from spatially smoothed

705   currents to indicate flow direction and qualitatively illustrate large-scale circulation

706   features, including surface divergence. Light gray areas have less than 12 drifter days per

707   square degree. In addition, only bins with mean current speeds statistically different from

708   zero at one standard error of the mean are shaded.

709

710   Figure 3. Variance ellipses centered every 5º longitude by 2º latitude (black lines) over

711   the square root of the magnitude of variance (colors in cm s^{-1}). Light gray areas have less

712   than 12 drifter days per square degree.

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714   Figure 4. Speed of currents regressed onto the Southern Oscillation Index (colors in cm s^{-}

715   ^{1}) shown for SOI = -1 (moderate El Niño) with directions (black arrows) indicated at for a

716   subset of gridpoints. Gridpoints with values that are not significantly different from zero

717   at one standard error of the mean not displayed.

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719   Figure 5. Near-surface ocean currents from surface drifter data. Details follow Fig. 2,

720   except displayed for climatological February (top panel) and August (bottom panel).

721

722   Figure 6. Near-surface ocean currents from surface drifter data. Details follow Fig. 2,

723   except displayed for climatological May (top panel) and November (bottom panel).

724

725   Figure 7. Seasonal variations of surface currents (black arrows, see key in lower left

726   panel) in the northwestern tropical Pacific Ocean. Currents are not shown where they are

727   not significantly different from zero at one standard error of the mean.

728

729   Figure 8. Seasonal cycle of surface divergence (colors, 10^{-8} s^{-1}) vs. latitude averaged over

730   longitude bands in the interior of the a) Indian (90ºE –55ºE), b) Pacific (165ºE–85ºW),

731   and c) Atlantic (30ºW–5ºE) Oceans. Tick marks show the start of named months.

732

733   Figure 9. Fractions (colors, upper colorbar) of (a) time mean, (b), eddy, and (c) seasonal

734   contributions to the total surface kinetic energy in the drifter climatology along with (d)

735   the total kinetic energy (colors, m^2 s^{-2}, lower colorbar).

736

737   Figure 10. Vector differences (arrows; colors indicate difference magnitudes in cm s^{-1}),

738   time-mean velocity of drifters minus a) January 1993–December 2003 time-mean total

739   speed from NOAA's 1/3 degree unsmoothed Ocean Surface Current Analysis Real-time

740   (OSCAR) product [*Bonjean and Lagerloef*, 2002] including geostrophic, Ekman and

741   buoyancy components, and b) time-mean geostrophic speed from the CNES-CLS 2009

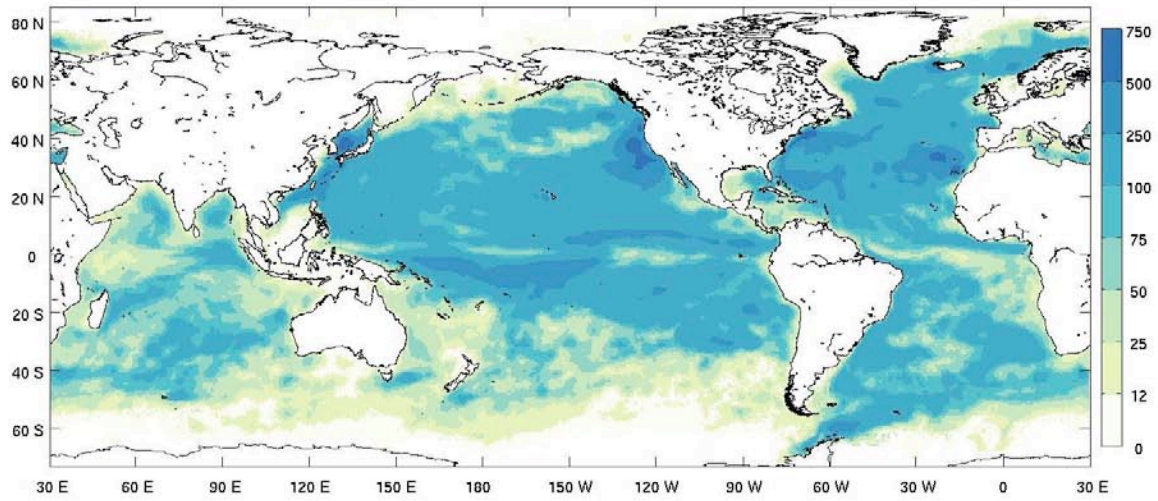742   v1.1 Mean Dynamic Topography [*Rio et al.*, 2011].

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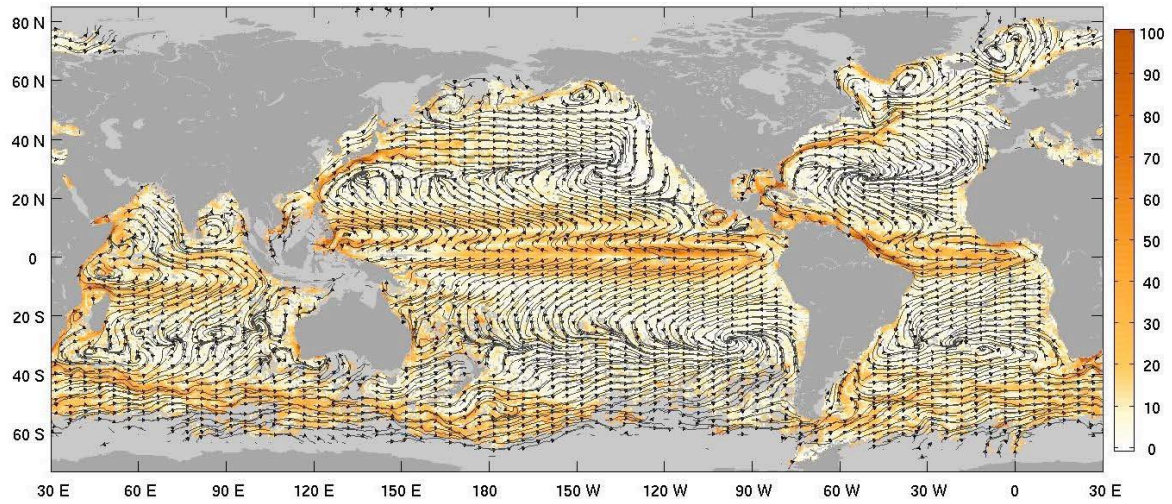bin (in drifter days per square degree).

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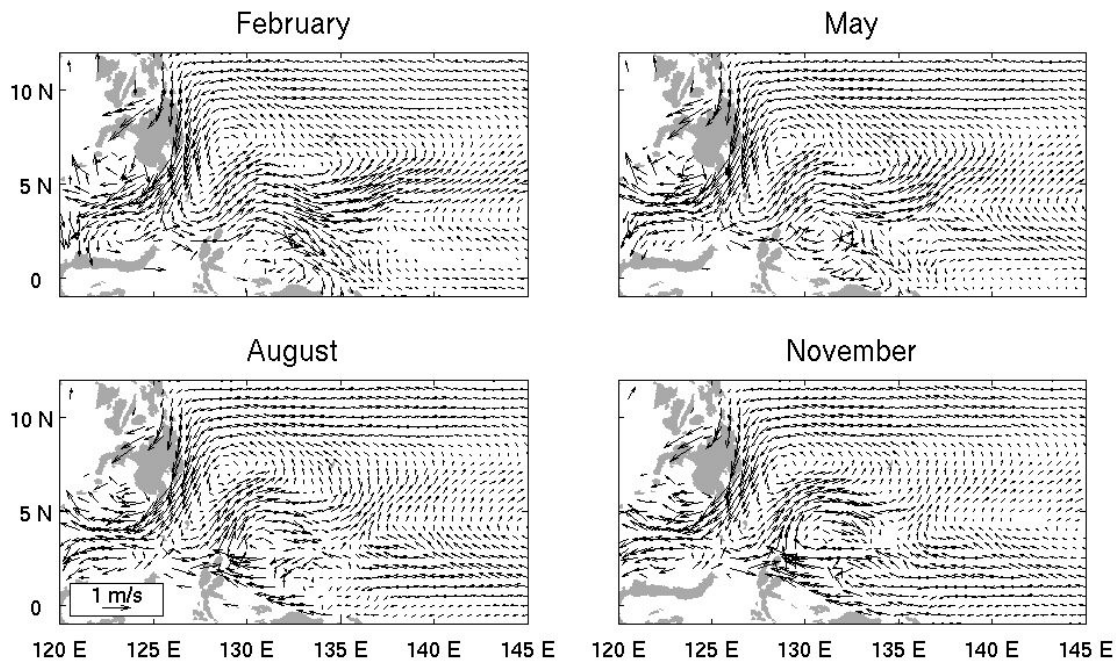
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February   May

August   November

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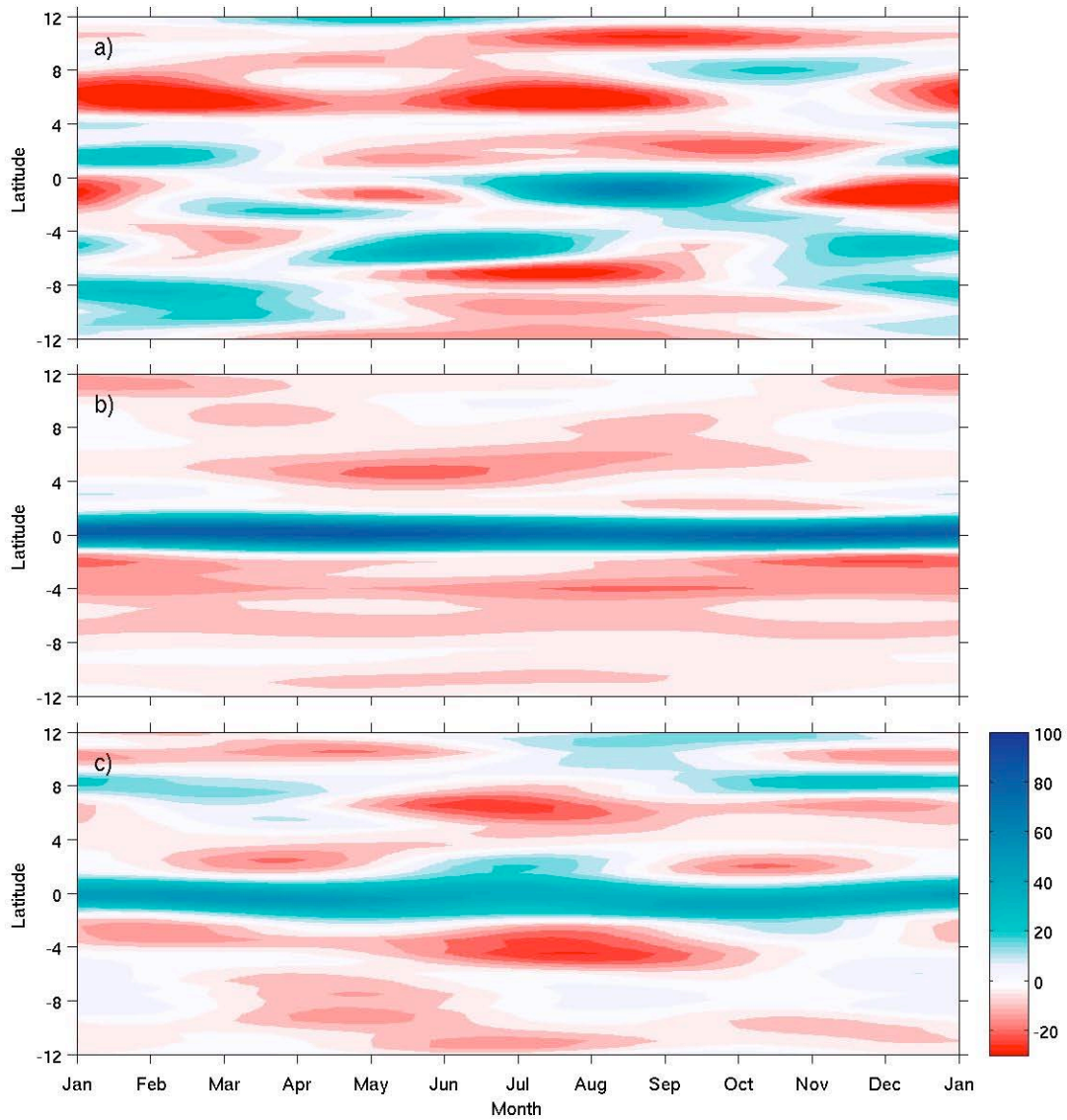
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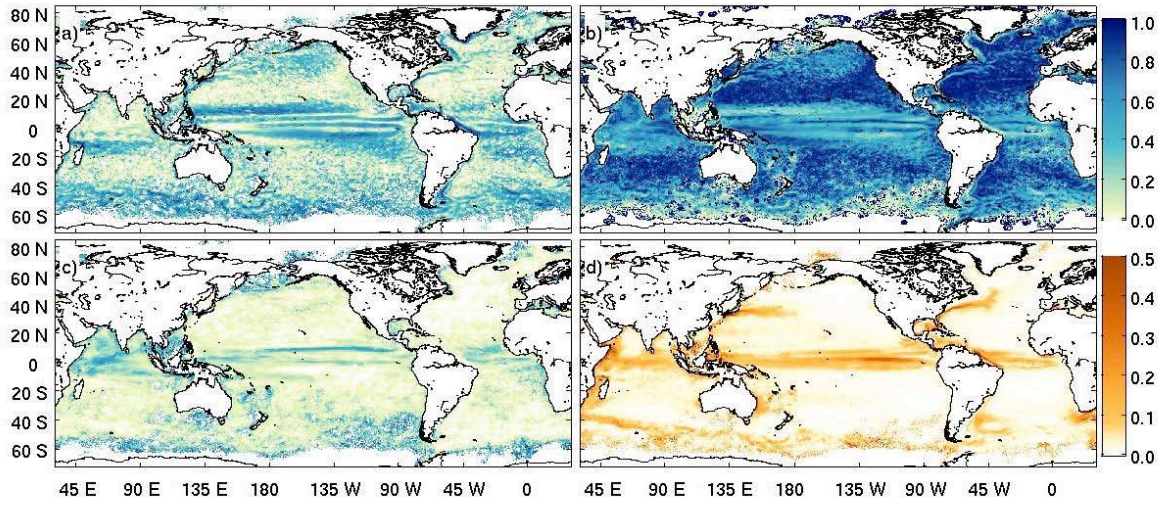
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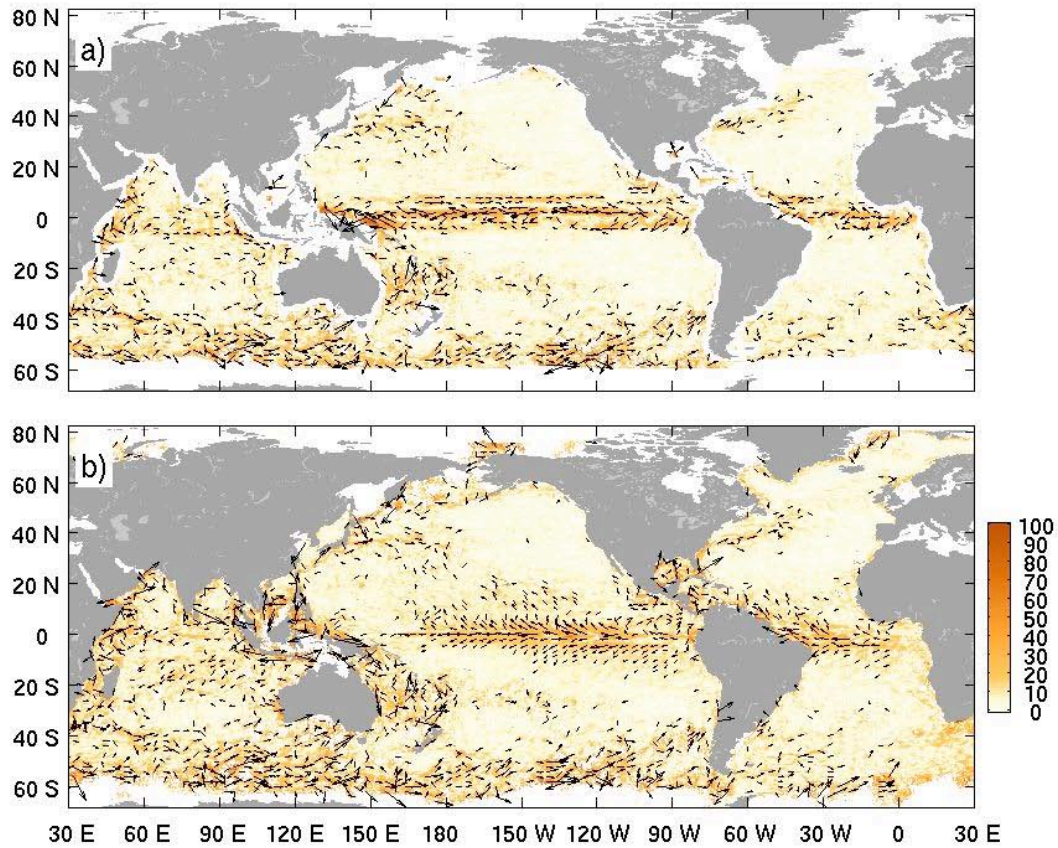
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